Satellite Retrieval of Lower-Tropospheric Ice Crystal Clouds in the Polar Regions

LYLE D. WILSON

Department of Meteorology, The Pennsylvania State University, University Park, Pennsylvania

JUDITH A. CURRY

Program in Atmospheric and Oceanic Sciences and Department of Aerospace Engineering Sciences, University of Colorado, Boulder, Colorado

THOMAS P. ACKERMAN

Department of Meteorology, The Pennsylvania State University, University Park, Pennsylvania

19 August 1992 and 1 December 1992

ABSTRACT

This paper investigates the satellite detection of low-level ice crystal clouds in the Arctic using a radiative transfer model and aircraft observations for six cases. Brightness temperatures at the top of the atmosphere are calculated for the six cases and for the corresponding clear-sky cases. Sensitivity studies are conducted to determine the sensitivity of brightness temperature at the top of the atmosphere to ice crystal layer optical depth, clear-sky brightness temperature, and height of the top of the ice crystal layer. These results are interpreted in the context of the ISCCP cloud retrieval algorithm.

1. Introduction

Schweiger and Key (1992) have recently compared a surface-based arctic cloud climatology with two satellite cloud climatologies. The surface-based climatology is that compiled by Warren et al. (1988), and the satellite climatologies are obtained from the International Satellite Cloud Climatology Project (ISCCP) (Rossow and Schiffer 1991) and from the *Nimbus-7* Global Cloud Climatology, hereafter referred to as C-MATRIX, of Stowe et al. (1988). In their comparison of the three climatologies, Schweiger and Key found that the disagreement between the surface-based and satellite climatologies over the central Arctic Ocean was quite large, particularly during winter, with the satellite climatologies reporting substantially more cloud than the surface-based climatology. For January poleward of the Canadian Archipelago, the Warren surface-based climatology gives an average of 45% cloud cover, the ISCCP dataset gives an average of 62% cloud cover, and the C-MATRIX dataset shows an average of 80% cloud cover. By contrast, summertime cloud amounts as determined by ISCCP are 20% less than those determined from the surface observations.

Because of the lack of visible and thermal contrast between cloud and the cold underlying surface in the polar regions, satellite detection of cloud in the Arctic has been very difficult (e.g., Key and Barry 1989; Raschke et al. 1992). However, the systematic overestimation of wintertime cloud by satellite when compared with the surface observations suggests that the satellite may actually be observing something that is missed by the surface observations. One possible explanation for the discrepancy is that the satellite is observing clear-sky ice crystal precipitation, which is difficult to observe from the surface and is not classified as cloud by surface observers.

The characteristics of the arctic clear-sky ice crystal precipitation, which are essentially those of a low-level ice crystal cloud, are summarized by Curry et al. (1990). Clear-sky ice crystal precipitation forms during the cold half of the year principally as a result of isobaric radiative cooling, and is observed to occur at atmospheric temperatures lower than $-15°$ to $-20°C$. An ice crystal "effective radius" of 40 μm and visible optical depths as large as 21 have been determined from the wintertime aircraft measurements.

Although the frequency of occurrence of these low-level ice crystals is uncertain, arguments are given by Curry et al. (1990), Curry and Ebert (1992), and Overland and Guest (1991) that these ice crystals are very widespread in the Arctic during the cold half of the year, with an estimated 50% frequency of occurrence during the winter season. Curry et al. (1990)

Corresponding author address: Dr. Judith A. Curry, Department of Aerospace Engineering Sciences, University of Colorado, Campus Box 429, Boulder, CO 80309.

carried out detailed radiative transfer calculations using observed vertical profiles of temperatures, humidity, and ice crystal concentrations and found perturbations to the incoming surface longwave flux associated with the ice crystals to be as large as 79 W m^{-2}. The observations and calculations of Overland and Guest (1991) also support the importance of low-level ice crystals to the surface radiation balance. The importance of the clear-sky ice crystal precipitation to climate arises principally due to the sensitivity of sea ice characteristics to the surface radiation flux (e.g., Ebert and Curry 1992b). Errors in the surface energy balance of the order of 80 W m^{-2}, or even of 10 W m^{-2}, have the potential to substantially modify calculations of sea ice characteristics and thus influence modeled global climate through the ice albedo feedback mechanism (e.g., Barry et al. 1984) and other radiation feedbacks between the atmosphere and sea ice as described by Ebert and Curry (1992b). Using a one-dimensional model of the coupled sea ice–atmosphere system, Curry and Ebert (1990) showed that the infrared heating associated with the clear-sky ice crystal precipitation dominates its impact on the surface radiation balance, and that simulations without the ice crystal precipitation resulted in an annually averaged equilibrium sea thickness of 4.1 m, compared with 2.95 m when ice crystal precipitation is included.

Because of its impact on the surface radiation balance, we believe that clear-sky ice crystal precipitation should be included in cloud climatologies. The question remains as to whether satellite cloud climatologies of the arctic region are interpreting the clear-sky ice crystal precipitation as cloud. Schweiger and Key (1992) addressed this issue by using mean January temperature and humidity profiles and an assumed ice crystal optical depth as input to a simple radiative transfer model. Assuming the top of the ice crystal layer to be near the top of the temperature inversion, they determined that brightness temperatures at the top of the atmosphere *increased* owing to the presence of the ice crystals. They concluded that it is unlikely that the ISCCP algorithm would detect clear-sky ice crystal precipitation as cloud, and that the discrepancy between the surface and satellite arctic winter cloud climatologies cannot be explained by satellite detection of clear-sky ice crystal precipitation.

In this paper we further investigate the possible effect of low-level ice crystals on satellite-based cloud statistics for the Arctic. Calculations of window brightness temperature at the top of the atmosphere are determined using a radiative transfer model for several observed cases of lower-tropospheric ice crystal clouds. For each case, calculations of brightness temperature for the observed cases are compared with clear-sky calculations. The difference between these two brightness temperatures are then interpreted in the context of the ISCCP cloud retrieval algorithm. Sensitivity tests are also conducted to assess the sensitivity of the window brightness temperature at the top of the atmosphere to various features of the ice crystal layers.

2. Radiative transfer models

The two-stream model outlined in Curry and Herman (1985) is employed for the infrared calculations, with some minor modifications, to calculate the upward intensity, I, at the top of the atmosphere. The two-stream equation for the upward intensity is

$$I_{\Delta\lambda} = B_{\Delta\lambda}(0)T_{\Delta\lambda}(z, 0) + \int B_{\Delta\lambda}(z')dT_{\Delta\lambda}(z, z'), \quad (1)$$

where $B_{\Delta\lambda}(z)$ is the integrated Planck function over the spectral region $\Delta\lambda$, and $T_{\Delta\lambda}(z, z')$ is the transmittance between heights z and z'. For the purpose of this study, $\Delta\lambda$ is taken to be 10.3–11.3 μm, corresponding to channel 4 of the Advanced Very High Resolution Radiometer (AVHRR) on board the NOAA polar orbiting satellites (from which the ISCCP cloud statistics are determined for the Arctic).

The relevant gaseous constituent of the model in this spectral region is water vapor, which is a weak absorber in this spectral region. The water vapor transmittance is calculated using the analytic transmission functions described by Kuo (1977). Kuo's transmission function for the spectral region of interest here accounts for continuum absorption, and a comparison with the transmission calculated as in Roberts et al. (1976) indicates that Kuo's method yields similar values. The absorption optical depth and transmittance for the ice crystals are determined following Ebert and Curry (1992a), which requires ice water content and ice crystal effective radius as inputs for the calculation. Determination of ice crystal microphysical properties for these cases is described by Curry et al. (1990). Total transmittance is the product of the water vapor and ice crystal transmittance. Brightness temperature is then obtained by inverting the Planck function in the 10.3–11.3-μm spectral region.

Calculations of visible reflectivities are made using a basic two-stream model (e.g., Liou 1974) with surface albedo added (see Cuzzi et al. 1982 for a four-stream example). Ice crystal shortwave optical properties (optical depth, single-scattering albedo, and asymmetry factor) are determined following Ebert and Curry (1992a).

Since the model results are to be interpreted in the context of the ISCCP cloud detection algorithm, the decision-making process for the algorithm that is used in the Arctic is summarized here. The ISCCP cloud dataset and algorithms are described in detail by Rossow and Schiffer (1991) and Rossow and Garder (1993). To determine the presence of cloud, the ISCCP algorithm uses a bispectral (visible and infrared) threshold when sunlight is present, and a single infrared threshold during the night. Over a snow/ice surface when sunlight is present, a pixel with either a brightness temperature at least 4 K colder than clear sky or a visible radiance at least 12% larger than clear sky is labeled as cloudy (Rossow and Garder 1992). Because of the lack of visible contrast between clouds and the underlying snow/ice surface, the ISCCP cloud algorithm is equivalent to an infrared-only method over sea ice (Rossow and Garder 1993). At night, only the infrared threshold is relevant. An additional distinction is made between cloudy pixels with radiances near the clear-sky values (brightness temperature difference between 4 and 8 K), which are referred to as "marginally cloudy," and cloudy pixels with radiances farther from the clear-sky values (brightness temperature difference > 8 K). If only one of these criteria is met, the pixel will be labeled as "marginally cloudy." Rossow and Garder (1993) believe that these "marginal clouds" are actually clouds, and Schweiger and Key (1992) found that with the marginal clouds included in the climatology, the satellite- and surface-based cloud climatologies are much closer in terms of cloud amount. However, the interpretation of the ISCCP marginal clouds must remain uncertain.

3. Data

Observations of the physical characteristics and vertical structure of the lower-tropospheric ice crystals are unfortunately sparse. Six profiles are used here (Table 1) that have been summarized by Curry et al. (1990). These profiles consist of three wintertime profiles that were reported by Witte (1968) and three springtime cases from the Arctic Gas and Aerosol Sampling Program (AGASP) (see Curry et al. 1990 and Meyer et al. 1991 for discussion), shown in Fig. 1.

The three wintertime profiles reported by Witte (1968) were all obtained from aircraft flights during December 1967 in the vicinity of Barrow, Alaska. Ice crystal layers as deep as 3400 m with average ice crystal concentrations as large as 1100 L^{-1} were measured using a Formvar replicator, although these concentrations may have been affected by fragmentation of the larger ice crystals. Since the ice crystal concentrations at the top and bottom aircraft traverses were significantly greater than zero, it can be inferred that the ice crystal layers extend beyond the levels reported (Curry et al. 1990).

TABLE 1. Case summaries and calculated infrared optical depths of the ice-crystal layers. "Top" and "bottom" refer to the top and bottom of the ice crystal layer.

Flight	Date	Top (m)	Bottom (m)	Optical depth	Ice crystal concentration (l^{-1})
A	2 December 1967	2900	150	2.7	450
E	9 December 1967	2260	550	3.3	500
H	13 December 1967	3400	280	10.7	1100
1084	7 April 1983	750	0	0.18	960
1090	18 April 1983	580	0	0.015	126
1240	13 April 1986	3100	950	1.9	3900

The three cases from AGASP were from obtained from aircraft flights during April of 1983 and 1986. The 1983 measurements were taken over the Beaufort–Chukchi Sea in the vicinity of Barrow, Alaska (Radke et al. 1984), while those in 1986 were taken over the eastern North American Arctic (Brock et al. 1990). The cloud physics instruments used included the Particle Measuring Systems FSSP, 1D precipitation probe, 2D probes, and an optical ice particle counter. Details of the profiles employed in this study are described by Curry et al. (1990) and Meyer et al. (1991). Average ice crystal concentrations for these cases range from 126 to over 4000 L^{-1}. However, Curry et al. (1990) cite a factor-of-2 uncertainty in these ice crystal concentrations that arises due to uncertainties in the assumed small-particle concentrations. The deepest ice crystal layer of the springtime cases is seen from Table 1 to extend in flight 1240 above 3100 m; unfortunately measurements above this level were not made.

4. Results

To assess the impact of the ice crystal layers on the infrared intensity received by the satellite, the model is run for all six cases both with and without the ice crystals present. As seen in Table 2, none of the three wintertime cases studied has brightness temperatures that are greater than 4 K colder than for the clear-sky case and only one of the springtime cases has a brightness temperature that is 4 K colder than for the clear-sky case. The albedos for the springtime cases were only 3%–4% brighter than the underlying snow albedo (which is assumed to be 82%; see Ebert and Curry 1992b). Therefore, only one of the six cases (flight 1240) would be labeled as cloudy according to the ISCCP algorithm. Flights A, E, and 1084 all have brightness temperatures colder than the clear-sky case, but the difference is less than the 4-K value necessary to receive a cloudy classification, although flight A only misses by 1 K. Flight 1090, with its shallow, optically thin ice crystal layer, has no discernible difference in brightness temperature between the clear sky and ice crystal cases. Flight H, on the other hand, has a brightness temperature that is 7.1 K *warmer* than the corresponding clear-sky brightness temperature, due to the very strong temperature inversion for this case (see Fig. 1). In light of the thresholds for clouds using the ISCCP algorithm, it is clear from Table 2 that only one of the six cases would be labeled as cloud, and that case would be classified as marginally cloudy.

Given the uncertainties in the aircraft observations of the ice crystal layers, most notably particle concentration and size and the top of the ice crystal layers,

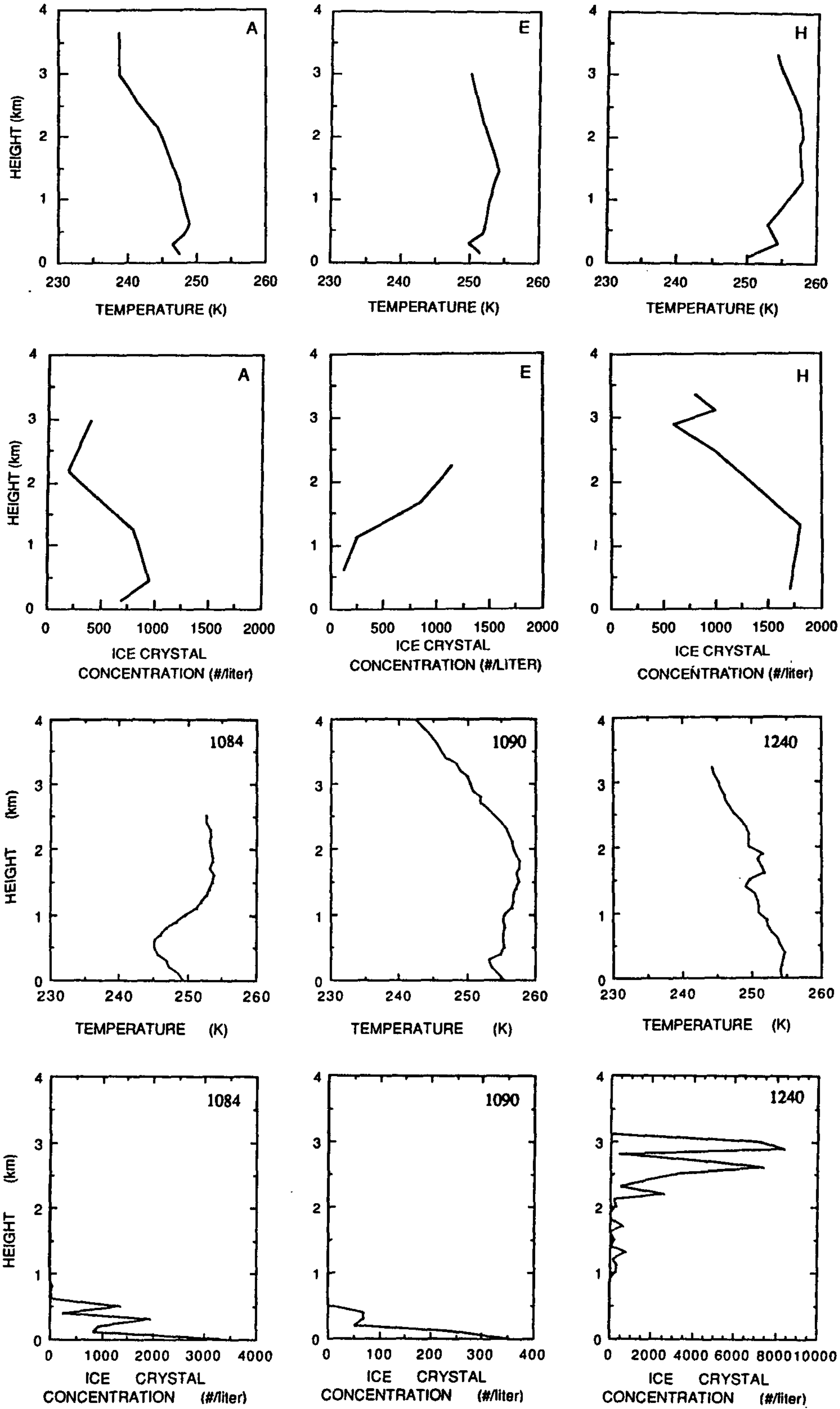

FIG. 1. Vertical profiles of temperature and ice crystal concentration for the six flights described in Table 1.

TABLE 2. Results from the clear-sky and ice crystal runs of the model for each case studied. Brightness temperatures that would be observed from a satellite are given as well as the difference of the ice crystal minus the clear-sky case.

Flight	Clear sky (K)	Ice crystal (K)	Difference (K)
A	248.6	245.6	−3.0
E	252.5	252.4	−0.1
H	248.1	255.2	7.1
1084	249.1	248.7	−0.4
1090	255.4	255.4	−0.0
1240	254.3	247.5	−6.8

and the uncertainties in the satellite retrievals of the clear-sky temperatures, it is instructive to examine the sensitivity of the brightness temperature at the top of the atmosphere to these parameters. Here we examine the sensitivity of the modeled brightness temperature at the top of the atmosphere to changes in ice crystal optical depth, the height of the top of the ice crystal layer, and clear-sky temperature.

To determine the effect of varying ice crystal infrared optical depth on the brightness temperature, the optical depth, τ, is varied for flights 1240 and H, with all other parameters held the same. As can be seen in Table 3, changes in τ do lead to changes in brightness temperature for both cases. This is not surprising, since the effective emission level changes for varying optical depth. The effect of varying τ on the observed brightness temperature therefore depends on the vertical temperature profile, which can be seen by comparing the two cases. For flight H, the brightness temperature increases with increasing optical depth at first, then decreases, consistent with the presence of a strong temperature inversion. For flight 1240, on the other hand, brightness temperature decreases with increasing optical depth since no significant inversion is present for this case. For the high surface albedos that are applicable here, the albedo at the top of the atmosphere shows little sensitivity to ice crystal optical depth.

To test the sensitivity of brightness temperature to height of the top of the ice crystal layer, the ice crystals in flights A and 1084 were extended upward, keeping τ the same. For flight A, raising the top of the ice crystal layer by 400 m is sufficient for the case to be labeled as cloudy. By extending the ice crystals in flight 1084 upward, the brightness temperature warms.

Errors in determination of the clear-sky brightness temperature could arise from contamination by ice crystals in "clear-sky" regions, errors in the "clear-sky" surface temperature, and the surface temperature in the clear-sky regions being different from that in the region where ice crystals are present. It is seen from Table 2 that an error of 1°C in clear-sky brightness temperature would result in a cloudy classification for flight A.

5. Summary and conclusions

Schweiger and Key (1992) tested their hypothesis that clear-sky ice crystal precipitation would not be detected as cloud by a satellite using the ISCCP algorithm. Using mean January temperature and humidity profiles and assuming the top of the ice crystal layer was near the top of the temperature inversion, they found that top of the atmosphere brightness temperatures over a cold surface were increased by the presence of the ice crystals and thus would not be classified as cloud. Based on their assumptions, this result is not surprising. However, results from the six cases presented here show that while the type of situation tested by Schweiger and Key does exist, other cases exist where these conditions are not met and brightness temperatures at the top of the atmosphere are lowered considerably by the presence of the ice crystals. Although only one of the six cases studied here had a brightness temperature decrease large enough to be labeled as cloudy by the ISCCP algorithm, sensitivity tests show that, depending on the ice crystals concentration relative to the shape of the vertical temperature profile, a small increase in optical depth or increase in the height of the ice crystals could result in a cloudy classification. Additionally, the cloud discrimination is shown to be sensitive to small errors in the clear-sky brightness temperature determination.

Unfortunately, not enough data is currently available to reach any firm conclusions regarding the satellite detection of low-level ice crystals in the Arctic. However, these results suggest that satellite detection of clear-sky ice crystal precipitation as cloud does occur using the ISCCP algorithm, although many cases probably remain undetected by the algorithm. This would suggest that at least some of the wintertime deficit in the cloud cover of surface-based as compared to satellite climatologies is because of the failure of the surface observations to include these ice crystals as cloud.

Additional data are needed to further our knowledge

TABLE 3. Sensitivity results showing modeled changes in satellite observed brightness temperatures as optical depth is multiplied by various factors. Note that flight H has a temperature inversion, while flight 1240 has no temperature inversion.

Flight	Mult. factor	Brightness temperature (K)
1240	0.1	253.1
	0.5	249.7
	1.0	247.5
	1.5	246.5
	3.0	245.5
H	0.1	253.7
	0.5	256.0
	1.0	255.2
	1.5	254.7
	3.0	254.2

of the role of clear-sky ice crystal precipitation in the radiation balance of the polar regions. However, results presented here and elsewhere suggest that the role played by these ice crystals in the arctic radiation budget is significant, and that these ice crystals should be included in future cloud climatologies.

Acknowledgments. Comments from W. Rossow, A. Schweiger, and the anonymous reviewers are appreciated. This research has been supported by Grants NSF DPP-8858830 and DPP-8820905.

REFERENCES

Barry, R. G., A. Henderson-Sellers, and K. P. Shine, 1984: Climate sensitivity and the marginal cryosphere. *Climate Processes and Climate Sensitivity, Geophysics Monograph* 29, J. E. Hansen and T. Takahashi, Eds., 221 pp.

Brock, C. A., L. F. Radke, and P. V. Hobbs, 1990: Sulfur in particles in Arctic hazes derived from airborne in situ and lidar measurements. *J. Geophys. Res.*, **95**, 22 369–22 387.

Curry, J. A., 1983: On the formation of continental polar air. *J. Atmos. Sci.*, **40**, 2279–2292.

——, and G. F. Herman, 1985: Infrared radiative properties of summertime Arctic stratus clouds. *J. Climate Appl. Meteor.*, **24**, 525–538.

——, and E. E. Ebert, 1990: Sensitivity of the thickness of Arctic sea ice to the optical properties of clouds. *Ann. Glaciol.*, **14**, 43–46.

——, and ——, 1992: Annual cycle of radiation fluxes over the Arctic Ocean: Sensitivity to cloud optical properties. *J. Climate*, **5**, 1267–1280.

——, F. G. Meyer, L. F. Radke, C. A. Brock, and E. E. Ebert, 1990: Occurrence and characteristics of lower tropospheric ice crystals in the Arctic. *Int. J. Climatol.*, **30**, 342–357.

Cuzzi, J. N., T. P. Ackerman, and L. C. Helmle, 1982: The delta-four-stream approximation for radiative flux transfer. *J. Atmos. Sci.*, **39**, 917–925.

Ebert, E. E., and J. A. Curry, 1992a: A parametrization of ice cloud optical properties for climate model. *J. Geophys. Res.*, **97**, 3831–3836.

——, and ——, 1992b: An intermediate one-dimensional thermodynamic sea ice model for investigating ice–atmosphere interactions. *J. Geophys. Res.*, in press.

Key, J., and R. G. Barry, 1989: Cloud cover analysis with Arctic AVHRR data, 1. cloud detection. *J. Geophys. Res.*, **94**, 18 521–18 535.

Kuo, H-L., 1977: Analytic infrared transmissivities of the atmosphere. *Beitr. Phys. Atmos.*, **50**, 331–349.

Liou, K-N., 1974: Analytic two-stream and four-stream solutions for radiative transfer. *J. Atmos. Sci.*, **31**, 1473–1475.

Meyer, F. G., J. A. Curry, C. A. Brock, and L. F. Radke, 1991: Springtime visibility in the Arctic. *J. Appl. Meteorol.*, **30**, 342–357.

Overland, J. E., and P. S. Guest, 1991: The Arctic snow and air temperature budget over sea ice during winter. *J. Geophys. Res.*, **96**, 4651–4662.

Radke, L. F., J. H. Lyons, D. A. Hegg, P. V. Hobbs, and I. H. Bailey, 1984: Airborne observations of Arctic aerosols. I: Characteristics of Arctic haze. *Geophys. Res. Lett.*, **11**, 393–396.

Raschke, E., P. Bauer, and H. J. Lutz, 1992: Remote sensing of clouds and surface radiation budget over polar regions. *Int. J. Remote Sens.*, **13**, 13–22.

Roberts, R. E., J. E. A. Selby, and L. M. Biberman, 1976: Infrared continuum absorption by atmospheric water vapor in the 8–12 mm window. *Appl. Opt.*, **15**, 2085–2090.

Rossow, W. B., and L. C. Gardner, 1993: Cloud detection using satellite measurements of infrared and visible radiances for ISCCP. *J. Climate*, in press.

——, and R. A. Schiffer, 1991: ISCCP cloud data products. *Bull. Amer. Meteor. Soc.*, **72**, 2–19.

Schweiger, A. J., and J. R. Key, 1992: Arctic cloudiness: Comparison of ISCCP-C2 and NIMBUS-7 satellite-derived cloud products with a surface-based cloud climatology. *J. Climate*, **5**, 1514–1527.

Stowe, L. L., C. G. Wellemeyer, T. F. Eck, H. Y. M. Yet, and the Nimbus-7 Cloud Data Processing Team, 1988: Nimbus-7 Global Cloud Climatology. Part 1: Algorithms and validation. *J. Climate*, **1**, 445–470.

Warren, S. G., C. J. Hahn, J. London, R. M. Chervin, and R. Jenne, 1988: Global distribution of total cloud cover and cloud type amounts over the ocean. NCAR Technical Note, TN-317+STR, 233 pp.

Witte, H. J., 1968: Airborne observation of cloud particles and infrared flux density (8–14 mm) in the Arctic. M. S. thesis, Department of Atmospheric Sciences, University of Washington 102 pp.